A STORM OVER THE PLATEAU, JANUARY 20-21, 1957

LEWIS C. NORTON AND MATTHEW H. KULAWIEC

National Weather Analysis Center, U. S. Weather Bureau, Washington, D. C.

1. INTRODUCTION

The large-amplitude ridge in the eastern Pacific Ocean (described by Stark [11] in the preceding article) caused most of the storms moving eastward across the western Pacific Ocean to be diverted northward over the Bering Sea or western Alaska during January 1957. This pattern also was favorable for cyclogenesis over the area to the northeast of the Hawaiian Islands. In this paper one of these storms is discussed from its arrival near the coast of Oregon until it reached the Central Plains States. Prior to arrival off the west coast, the storm had followed a circuitous path across the eastern Pacific Ocean. As the storm stagnated near the mouth of the Columbia River, its energy caused cyclogenesis which moved eastward across southern Oregon, southern Idaho, and Wyoming, reaching northeastern Colorado during the 24-hour period. An attempt will be made to describe the weather and changes associated with the movement of this storm.

2. ANTECEDENT CONDITIONS

The week preceding the arrival of the storm off the Pacific Coast was characterized by a long-wave trough at the 500-mb. level persisting near 70° W. as shown on both the Hovmöller [6] diagram and the Fjørtoft [4] mean flow charts. During most of the same period long-wave ridges had persisted near 20° W. and 150° W. This unusually long wavelength prevailed at higher latitudes over all of the Northern Hemisphere during this period. At lower latitudes a much shorter stationary wavelength existed because the mean zonal wind speed was not of sufficient magnitude to support the wavelength which prevailed at the higher latitudes.

A check of the mean zonal wind speed as derived from the Rossby [9] long-wave formula showed a value of about 62 knots for a stationary wavelength of 130° longitude at 50° N. If the refinements suggested by Haurwitz [5] are included in the computations the value of the zonal wind speed necessary for equilibrium is reduced to about 55 knots. These derived values were compared to the average westerly component as measured on the Fjørtoft space mean flow charts. The measured values averaged near 50 knots, with values exceeding 100 knots in the vicinity of the long-wave trough over southeastern Canada.

In association with a 500-mb. lower-latitude trough that persisted over the eastern Pacific Ocean during this same period, two surface Lows developed and moved northeastward toward the Oregon–Washington coast. Mild maritime air in advance of these storms maintained above normal temperatures over the Pacific Coast States and southern Plateau region. In portions of Arizona and New Mexico temperatures were as much as 12° F. above normal. In contrast, the strong northwesterly flow over the Northern Plains States and the northeastern portion of the United States maintained below normal readings with some departures over South Dakota exceeding 15° F. [13].

3. STORM DEVELOPMENT

PRIOR TO JANUARY 20

On January 14, a new Low center began to develop northeast of the Hawaiian Islands about 600 miles west of the preceding two storms. This storm moved north-northeastward for about 48 hours with only minor intensification. During the next 18 hours the Low swerved westward and deepened as it came more nearly under the influence of an intense Low aloft. By January 17, the surface Low had recurved northward as the circulation aloft moved northward around the west side of the blocking High in the Gulf of Alaska. Intensification of the ridge to the west of the Low aloft forced it and the accompanying surface system eastward toward the west coast of North America. During the northward movement, the surface Low underwent cyclolosis in sympathy with the weakened cyclonic circulation aloft.

Figure 1 shows the development of the trough aloft as it moved from near 150° W. on January 17 to near 120° W. on January 20. The associated weakening of the ridge to the east of this trough and the intensification of the one to the west should be noted.

The surface Low underwent cyclogenesis as it moved eastward from Ship Papa (50° N., 145° W.). This was due in part to the advection of cyclonic vorticity aloft into the vicinity of the storm and "digging" as described by Wobus [16]. On the 500-mb. chart for 0300 GMT, January 19, a current of 50-knot winds moving out of Alaska toward a col necessitated a crossing of the winds

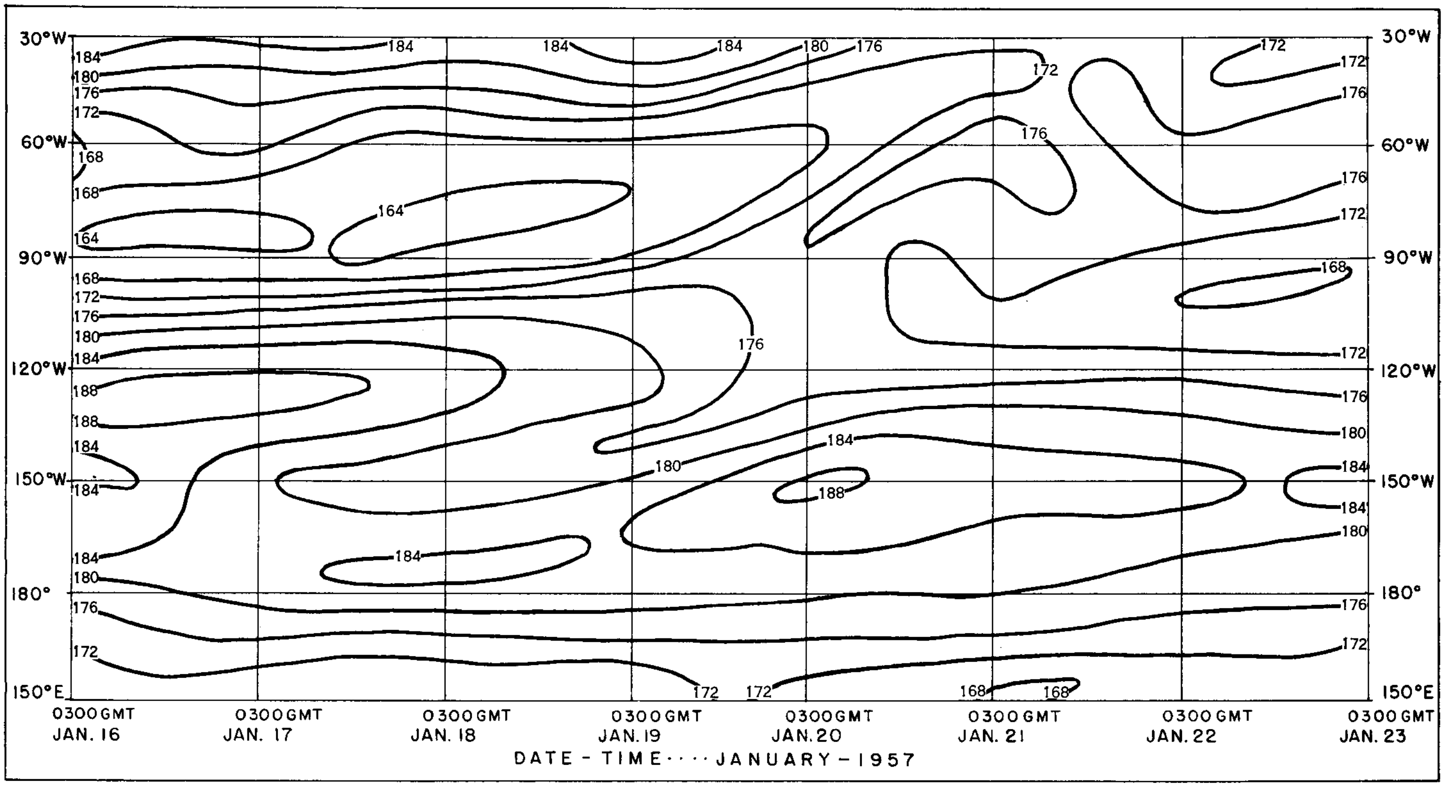

FIGURE 1.—Hovmöller diagram representing a cross section of the 500-mb. constant pressure chart from 30° W. to 150° E. during the period 0300 GMT, January 16, to 0300 GMT, January 23, 1957. Height values are in hundreds of feet.

toward higher contours and resultant slowing until they were in equilibrium with the actual gradient.

As the surface storm deepened and moved steadily toward the Oregon Coast, it advected increasing amounts of moisture over the Pacific Northwest. By 1230 GMT, January 19, the overcast preceding the storm had overspread nearly all of Oregon and Washington. Within another six hours precipitation had begun, primarily as snow, over western Oregon and Washington. South of the latitude of the storm center the winds in advance of the Low soon effected sufficient warming to change the precipitation to rain.

Figure 2 depicts the mean isotherm pattern which existed at 0300 GMT, January 19 prior to the arrival of the storm at the Pacific Coast. At that time a small 200-foot negative departure from normal in the thickness of the 1000–500-mb. layer indicated the extent of the colder air accompanying the storm center over the eastern Pacific Ocean. Meanwhile, the larger negative anomaly that was to play a prominent role in the development of this storm and the subsequent readjustment of the broad-scale features was located over northwestern Canada. The huge negative anomaly, which covered the area east of the Mississippi River at this time, marked the remnant of a retreating surge of cold air which had persisted over much of the eastern portion of the United States during the preceding weeks of January. The positive departure over Alberta and Saskatchewan was to be strengthened as warm advection ahead of the Pacific Low moved northeastward and tended to overspread the existing anomaly. It was this push of warmer air which later was to move eastward and produce mild temperatures in Washington, D. C., for the Presidential inauguration and a January thaw over the New England States.

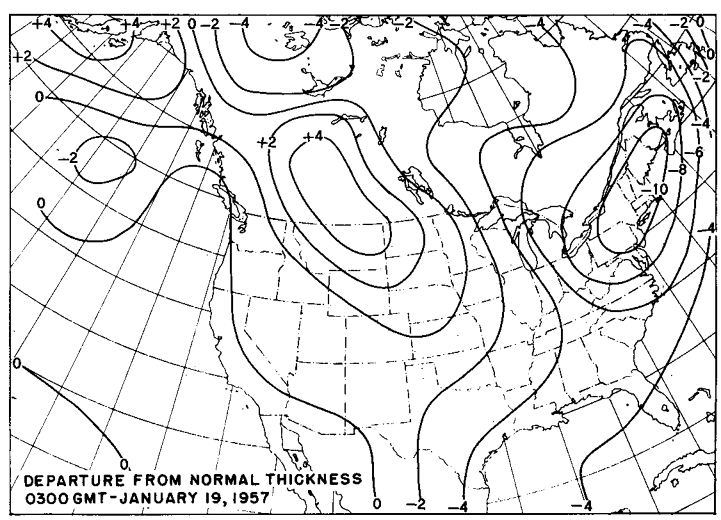

FIGURE 2.—The departure from normal of the mean virtual temperature of the 1000–500-mb. layer for 0300 GMT, January 19, 1957, as expressed in hundreds of feet of thickness.

JANUARY 20–21

By 0300 GMT, January 20 (fig. 3A) the surface Low advancing toward the northwestern coast of Oregon had approached within about 120 miles of the mouth of the Columbia River and had begun to decelerate. At the same time the precipitation had spread eastward across most of Idaho and southward into northern California. The continued warm advection to the southeast of the storm center had raised the freezing level over Medford to nearly 4,500 ft. m. s. l. thus changing most of the precipitation west of the Cascade range from snow to rain. However, over eastern Oregon most of the moisture fell in the form of snow as sufficient cold air remained from the previous stagnant polar air mass to maintain subfreezing temperatures at all levels. Precipitation in the area to the north of the storm center continued predominantly as snow, as the southeasterly and easterly winds were bringing colder air at lower levels from eastern Washington and southwestern Canada across the coastal plain of Washington. Precipitation falling through this drier continental air was evaporating some as it raised the moisture content of the airmass. This evaporation contributed to the cooling of the airmass at lower levels and maintained temperatures below freezing through the troposphere.

Despite the fact that the major concentration of cyclonic vorticity aloft continued to be accompanying the parent surface Low (see fig. 3B) a sufficient amount of thermal anticyclonic vorticity existed over southwestern Oregon to favor cyclogenesis. A new storm center is shown in figure 3A over the upper Willamette Valley, although the data were not sufficient to define a closed circulation. The region where the apparent cyclogenesis occurred was favorable for trough development because the southwesterly winds aloft were moving downslope on the east side of the coastal ranges. This type of redevelopment inland has been noticed frequently by the authors as major storms approach the Pacific Coast. Some of the other mechanisms responsible for this "center jump" include the damming effect of the coastal ranges, the downslope effect producing a continued cyclogenetic region as winds moved out of eastern Washington and southern Canada toward the parent storm center, the fact that the offshore region is also a climatologically favorable long-wave trough position, and the existence of warmer air over the water favoring the maintenance of a Low in that region rather than over the colder winter land mass.

During the next 12 hours the original or parent Low continued to fill and the new surface storm center appeared (fig. 4A) as a definite closed circulation over the upper Snake River Valley in southern Idaho. Some deepening of the Low occurred during this period as colder air from northern Idaho and eastern Washington was advected into the rear of the Low. This advection was also permitted by the fact that the new Low had separated farther from the parent Low. Precipitation continued to spread eastward to central Wyoming and southward through California to the southern end of the San Joaquin Valley. The cloudiness ahead of the storm advanced to produce overcast conditions 300 to 400 miles farther east as sufficient moisture had been carried by the westerly

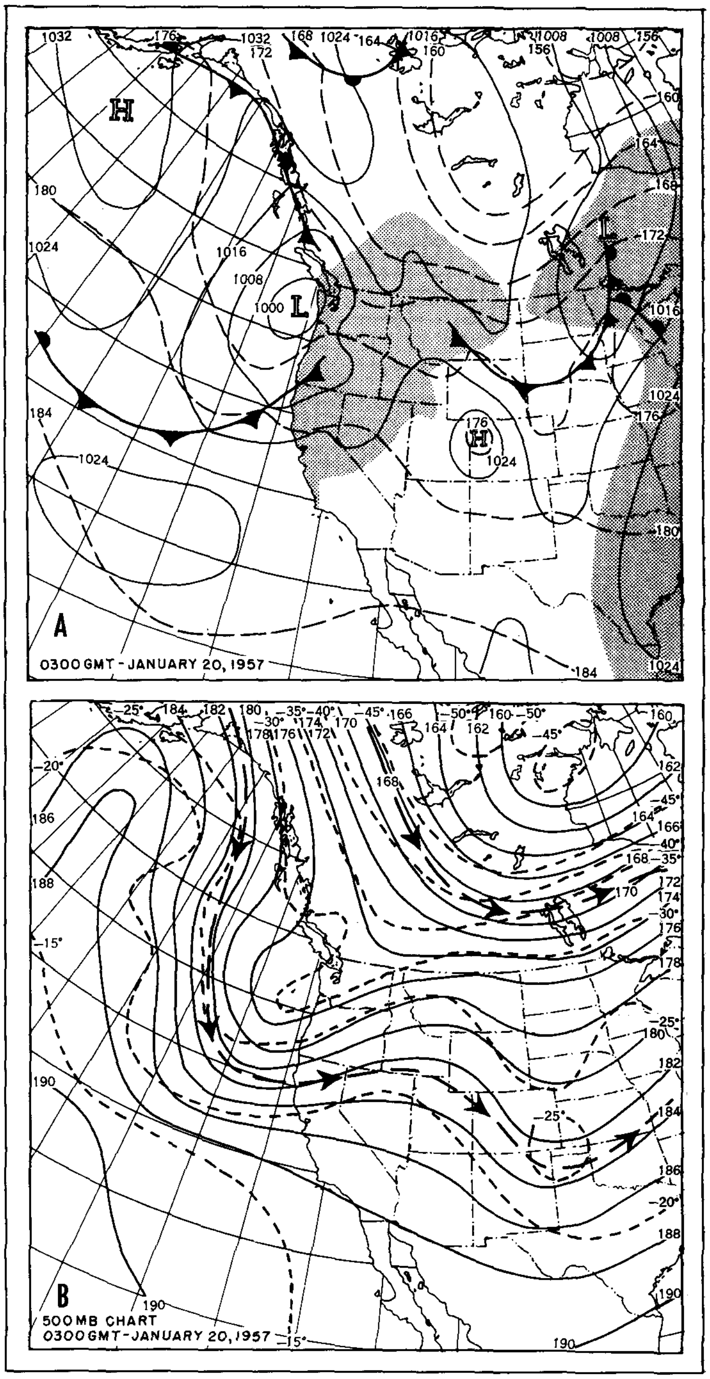

FIGURE 3.—Synoptic patterns for 0300 GMT January 20, 1957. (A) Surface chart (solid lines) with 1000–500-mb. thickness (dashed lines). Shaded area is essentially overcast. (B) 500-mb. chart with height contours (solid lines) and isotherms (dashed lines). Heavy lines with arrows indicate the position of the 300-mb. jet.

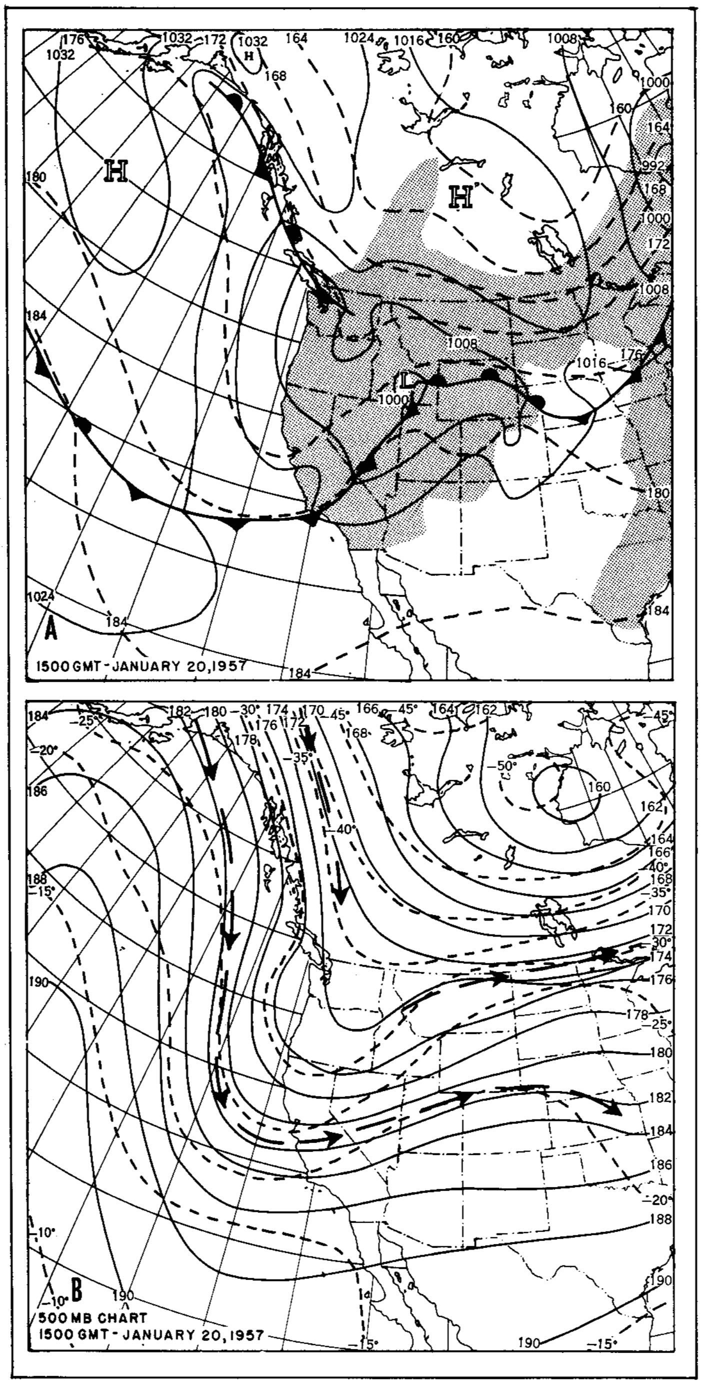

FIGURE 4.—Synoptic patterns for 1500 GMT, January 20, 1957. (A) Surface chart (solid lines) with 1000–500-mb. thickness (dashed lines). Shaded area is essentially overcast. (B) 500-mb. chart with height contours (solid lines) and isotherms (dashed lines). Heavy lines with arrows indicate the position of the 300-mb. jet.

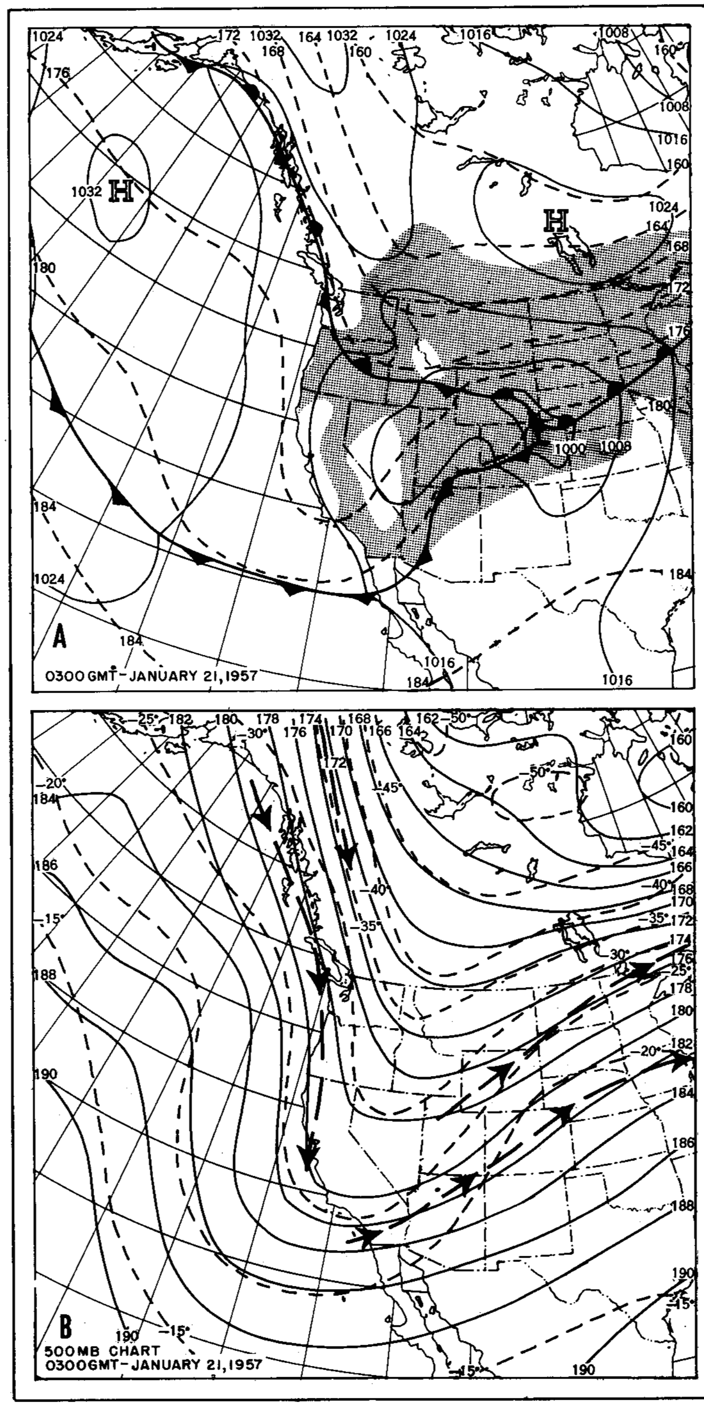

FIGURE 5.—Synoptic patterns for 0300 GMT, January 21, 1957. (A) Surface chart (solid lines) with 1000–500-mb. thickness (dashed lines). Shaded area is essentially overcast. (B) 500-mb. chart with height contours (solid lines) and isotherms (dashed lines). Heavy lines with arrows indicate the position of the 300-mb. jet.

winds aloft as far as southwestern Arizona and northeastern Colorado.

Aloft, the short-wave pattern had advanced eastward with the area of maximum cyclonic vorticity (fig. 4B) now located near the Idaho-Oregon border. During the same period, the northwest winds over the Rocky Mountains, which produced very little downslope conditions over eastern Colorado, backed to the west-southwest, favoring the intensification of the thermal trough immediately east of the Rockies as shown in figure 4A. The area of strongest difluence indicated on the 0300 GMT chart (fig. 3) over southern British Columbia had shown

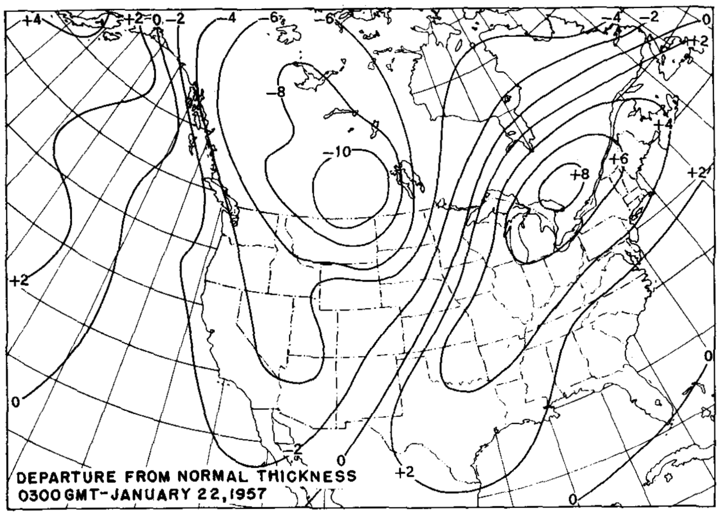

FIGURE 6.—The departure from normal of the mean virtual temperature of the 1000–500-mb. layer for 0300 GMT, January 22, 1957, as expressed in hundreds of feet of thickness.

slight movement. This slow movement was partially the result of a short-wave trough moving in the northwest flow toward the region of diverging contours. As it moved into the difluent zone this stream of air was being turned toward the right and lifted, as the wind pattern attempted to reach a balance with the existing gradient at the 500-mb. surface. This mechanism favored pressure falls over eastern Washington and pressure rises over the eastern Pacific to the right of the stream. This lifting was, in part, responsible for cooling observed over the Pacific Northwest as depicted by the increase in the size of the area enclosed by the −30° C. isotherm at the 500-mb. level over Oregon and Washington.

With the passage of another 12 hours (see fig. 5A) the storm center migrated eastward to the vicinity of Denver. It moved somewhat southeastward as it dropped into the thermal trough which had continued to intensify over eastern Colorado. Much of the moisture, originally available in the maritime airmass as it entered the Pacific Coast States the preceding day, had fallen in the form of rain and snow over the Plateau region. The moisture decrease was evident in the smaller areal extent of both the cloudiness and precipitation, although some snow showers continued in and along the western slopes of the central Rocky Mountain area. At this time the leading edge of the overcast preceded the frontal zone and associated precipitation by only 100 to 200 miles.

The difluent region on the 500-mb. chart over British Columbia at 1500 GMT had become much less prominent by 0300 GMT, January 21 (fig. 5B) due to the movement of the short wave in that current eastward into the Alberta area. Meanwhile the secondary area of marked difluence that existed in the vicinity of Vancouver Island (fig. 4B) had moved southward and intensified until it was the primary delta zone on the 500-mb. chart for 0300 GMT, January 21. In association with this difluent area a trough over Nevada had remained, despite the displacement of the frontal boundary eastward to southeastern Utah. The cold air moved southward behind the advancing front and by 0300 GMT, January 21, the −25° C. isotherm, representing a 24-hr. cooling of 10° C., reached into southern California.

By 0300 GMT, January 21, the portion of the 500-mb. chart at middle latitudes from the Central Plains westward to the eastern coast of Siberia (fig. 1) was in apparent equilibrium as evidenced by only slight fluctuations in the mean positions of the troughs and ridges during the following week. Some adjustment continued over the eastern half of North America and the North Atlantic Ocean. In this region the ridge over Quebec migrated slowly eastward and was replaced with a trough by the end of the week. This eastern trough, however, was confined primarily to the area north of 40° N. as the stronger zonal flow over the United States prevented the development of a full-latitude trough.

The positions of the mean trough and ridge on January 21 over the Northern Hemisphere were compared to the "Martin Anomaly Charts" [12]. Since the anomaly of greatest magnitude was located over the blocking High in the Gulf of Alaska, it was chosen as the key area or "anchor position" for our comparison. It showed that the other anomaly areas existing over the Northern Hemisphere were in agreement with the most favored positions.

To illustrate the magnitude of the broad-scale readjustments, the temperature changes that occurred over central North America may be examined by comparing anomaly charts preceding (fig. 2) and following (fig. 6) the period we have discussed. The large negative departure previously existing over the New England area was replaced by a positive anomaly. The small negative departures indicated over northern Canada and the eastern Pacific Ocean in figure 2 united and increased in magnitude as a result of the rapid development near the Pacific Coast and replaced the near or slightly above normal values previously covering the area.

This storm and the accompanying circulation changes were associated with a marked change in the position and intensity of the jet stream. As the storm approached the Pacific Coast on January 20, the primary jet stream was located from northeastern Alaska across northeastern Alberta and southern James Bay to the Canadian Maritime Provinces, then northeastward across the North Atlantic (fig. 3B). At the same time a secondary jet stream was moving out of southern Alaska southward across northern California. This latter branch coursed mainly eastward at mid-latitudes across the United States except for some southward displacement over the Central Plains. During the next 24 hours the increasing northerly flow off the Pacific Coast was accompanied by a westward shift of the jet stream across western Canada, while during the same interval the northerly jet stream over the ocean was displaced toward the coastal regions (fig. 4B). The shift in the position of the stream over the Plateau region was even more marked as the other jet stream, now the primary jet, migrated to extreme southern California (fig. 5B). This readjustment over the western half of the United States produced marked changes over the eastern United States and eastern Canada, with the jet stream shifting northward about 400 miles in that area.

VERTICAL DEVELOPMENT, JANUARY 20–21

The preceding discussion of events associated with this storm has been concerned mainly with horizontal charts. The very fortunate coincidence of the surface positions of the Low occurring very close to raob stations at the synoptic times of upper-air observations suggested a brief investigation of the vertical changes accompanying this storm.

The center of the surface Low was chosen for a point of reference. A vertical time cross section (fig. 7) was prepared, using for the ordinate the logarithm of the pressure and for the abcissa, time. Each 12-hr. interval is represented by a sounding from a station located very near to the center of the surface Low. The soundings for Boise, Idaho and Denver, Colo. are the only ones with important displacements from the center of the Low. Boise, at 1500 GMT, January 20, was located approximately 80 miles to the northwest of the storm center and Denver, at 0300 GMT, January 21, was approximately 40 miles west of the storm center. Although the previous investigation had considered the storm only during its movement from the Pacific Coast to the vicinity of Denver, Colo., certain changes shown by the cross section required its extension beyond Denver for another 24 hours.

The two parameters that lend themselves to a concise portrayal of vertical columns of air are the D values [1] and the free air temperatures. The D method is more suitable to portray pressure, which is a quantity with a dominant gradient in the vertical rather than in the horizontal [10]. The reference base of the D values is the National Advisory Committee for Aeronautics (NACA) atmosphere.

Along any isobaric surface, the minima in D values are the troughs in the plane of the cross section (vertical) and the maxima are the ridges. Stability conditions are indicated by the rate of increase or decrease of D values. Warming, then, is indicated by decreasing negative departures (or increasing positive departures), and, conversely, cooling by increasing negative departures (or decreasing positive departures). Because there is a certain relationship between D values and the mean virtual temperature, the maximum D values in the positive sense indicate air of lower density and, conversely, maximum D values in the negative sense air of higher density. Free air temperatures are included in the cross section because they are frequently used as an index of horizontal advective changes.

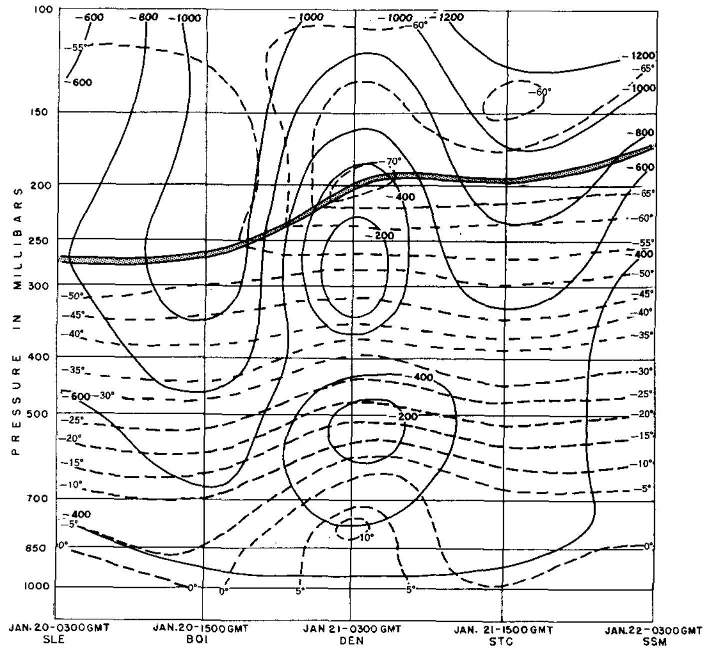

FIGURE 7.—Time cross section through the approximate center of the surface low. D values (solid lines) are drawn for every 200 feet. Isotherms (dashed lines) are drawn for every 5° C. The tropopause is shown as a double dot shaded line. SLE=Salem, Oreg., BOI=Boise, DEN=Denver, STC=St. Cloud, and SSM=Sault Ste. Marie.

Vederman [15] made an investigation of the changes of the vertical column of air above rapidly deepening storms. Unlike his cases, this particular storm exhibited no net deepening in its 48-hour movement from western Oregon to Sault Ste. Marie, Mich. Fluctuations were noted in the height of the 200-mb. surface, but for practical purposes the height of the 1000-mb. level can be assumed nearly constant.

Examination of this cross section (fig. 7) reveals a number of interesting features which certainly invite further study. In the vertical there is evident a 12-hour cycling pattern of rise and fall in D values. These changes, for the most part, are confined to the portion of the atmosphere above the 400-mb. level. The strong warming maximum at the 500-mb. level in the Denver sounding may be accounted for mostly by downslope warming. But it is interesting to note that alternate cooling and warming occurs even at lower levels of the column although the changes are not so prominent as above the 400-mb. level. One inference which suggests itself is the diurnal effect. It may be, perhaps, that alternate advective cooling and warming occurred during the motion of the storm, but the rhythmic nature of this event makes this explanation difficult to accept. No attempt has been made to determine whether those density changes were due to vertical motion. On the isobaric surfaces, the maximum changes occurred between the 300-mb. and 150-mb. levels with perhaps the greatest change at 250 mb.

In terms of the orographic effect, the changes in the air column as observed and the changes that can be calculated leave much to be desired in terms of accuracy at some places. The airmass in the lowest 5,000 feet was near saturation as it entered the Pacific Coast States. The moist adiabatic process is assumed when air is forced over mountain ranges. A value of 5,000 feet was assigned to the mean elevation of the Cascade Range. Taking the sounding at Salem, Oreg., and lifting the lower 10,000 feet of the column by 5,000 feet, gives a calculated drop of roughly about 5° C. in the lower 5,000 feet of the column and about 2° to 3° C. in the upper 5,000 feet.

With the heterogeneous topography existing between the Cascades and Boise, Idaho, it is difficult to ascribe accurately topographic effects beyond the Cascades. Here it is assumed that all other things being equal, a 5,000-foot lift would give a rough index of the change in the column due to lifting. The Boise sounding agreed reasonably well with these calculations. A similar calculation for the further lifting and downslope heating of the air reaching Denver also resulted in reasonable agreement with the observed sounding.

One could argue from the foregoing discussion that the cyclical temperature changes in the layer below the 400-mb. level at Boise and Denver could be accounted for by the topographic effects. These effects would not account for the changes at Saint Cloud, Minn., and Sault Ste. Marie, although the cyclical changes were of smaller magnitude in these latter soundings.

While the D values change on the isobaric surfaces quite markedly, the free air temperature changes on the same surfaces are comparatively small. For example, at the 250-mb. level the maximum temperature change is 6° C., whereas the maximum D value change is 1,000 feet. A 6° C. change in the mean virtual temperature in the layer between 300 mb. and 250 mb. would result in a thickness change of slightly over 100 feet. Similarly a 15° C. change in temperature throughout the atmosphere between 250 mb. and 200 mb. would give a thickness change of less than 350 feet. This indicates that while the changes between adjacent isobaric levels were quite small, the vertical changes of large portions of the column were striking.

No attempt was made to find a physical basis for these changes and features. A closer study would probably reveal other relationships but neither time nor space allow their recognition in an article of this scope.

4. THE STORM TRACK

A comparison was made between the track of this storm and the published normal tracks of Bowie and Weightman [2], the California Institute of Technology [3] and Klein [8]. Following the system of [2] the storm discussed here would be classed as a North Pacific storm since it reached the Pacific Coast north of 40° N. A storm of this type normally shows an east-northeasterly track until it reaches the longitude of eastern Oregon, then turns toward the southeast until it reaches the Central Plains and recurves toward the Great Lakes area. The normal storm shows an average movement of about 900 miles in each 24-hour period. The path of the storm of January 20–21, 1957, agreed well both in direction and speed with these average qualities.

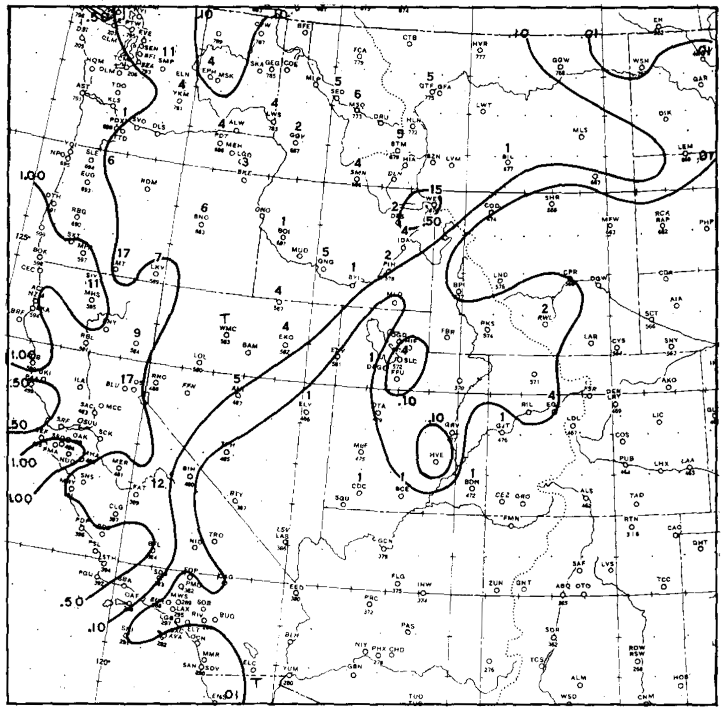

FIGURE 8.—Approximate precipitation totals for the period 1230 GMT, January 19, to 0630 GMT, January 21. Isohyets are drawn for 1.00 inch, 0.50 inch, 0.10 inch, 0.01 inch. Superimposed numerals indicate snow cover increase associated with the storm for the period ending 1230 GMT, January 21, 1957.

The January 1957 storm shows some similarities to the CIT type E storm although that is primarily a type of storm that occurs with stronger zonal conditions. The average storm tracks of Klein likewise show fair agreement with the actual track of this storm, although it must be remembered that these tracks of Klein represent composite tracks of many different types of storms as related to origin and direction of movement.

It might be expected that such a nearly "normal" storm would present little difficulty in the prediction of its movement and changes. This was borne out by a check of the verification scores of the surface prognostic charts prepared by the National Weather Analysis Center (NAWAC) to verify at 0030 and 1230 GMT, January 20 and 0030 GMT, January 21. Scores were better than the average for January 1957 and better than the 7-year mean for January. A check of the prognostic charts prepared by NAWAC and the thermotropic prognostic chart prepared by the Joint Numerical Weather Prediction Unit (JNWP) for the 500-mb. surface showed the same situation although the charts issued by JNWP were slightly better than those of NAWAC.

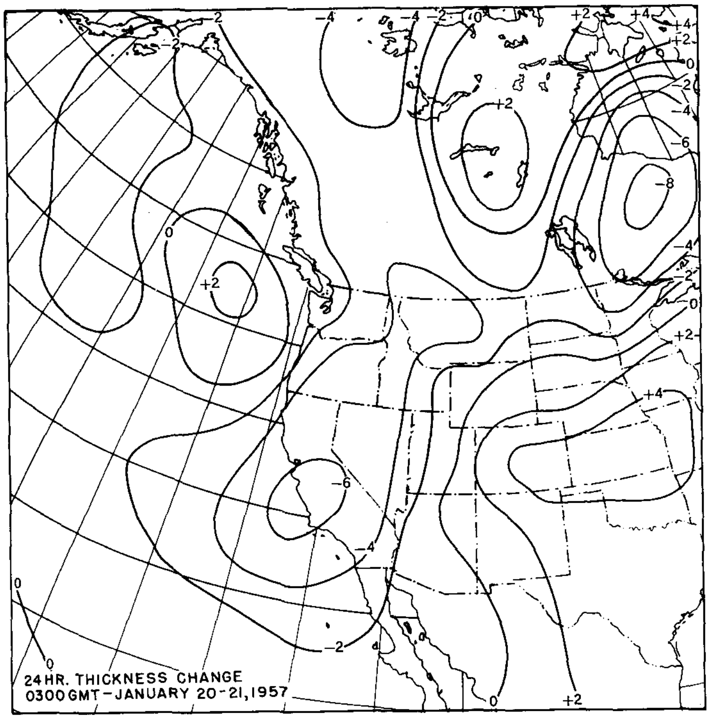

FIGURE 9.—Thickness change between 0300 GMT, January 20, and 0300 GMT, January 21, 1957, of the 1000–500-mb. layer in hundreds of feet.

5. PRECIPITATION AND TEMPERATURE

This storm brought rather widespread precipitation across much of the western Plateau but the heaviest amounts were confined to the immediate coastal areas south of the storm center. As shown in figure 8, the maximum amounts reported from this storm fell in northern California and southwestern Oregon, while a secondary maximum of over one inch was reported in the San Francisco Bay area. At most reporting points over 75 percent of the total precipitation received from the storm fell during an 18-hour period accompanying the frontal passage. The precipitation totals from this Low represented over the Plateau area about ⅓ to ½ of the total moisture received during the month [14]. Since a large portion of the precipitation at the higher elevations in the Sierra Nevada and Cascade ranges and to the east was in the form of snow, a substantial increase in the snow cover resulted. Some of the more significant increases reported included 17 inches at Crater Lake, Oreg., and Blue Canyon, Calif., 5 inches at Austin, Nev., and over 15 inches at West Yellowstone, Mont.

The initial surface temperature changes were not very spectacular since the new maritime airmass was not appreciably different in the lower levels from the existing stagnant continental airmass over the Plateau. Arctic air moved southward over that region during the following week and brought very sharp temperature drops with some all-time records being broken. The change in the thickness of the 1000–500-mb. layer is shown in figure 9 with a decrease of more than 600 feet (cooling) occurring over central California. Showalter (c. f. [7]) has pointed out that a 200-foot change in the 1000–700-mb. thickness is approximately equal to a change of 10° F. in the maximum surface temperature if the lapse rate remains constant. He has also shown that $Z_5=2Z_7-2F$, where Z_5 is the 1000–500-mb. thickness and Z_7 the 1000–700-mb. thickness and F is a measure of the stability of an airmass. Thus

$$\frac{\partial Z_5}{\partial t}=2\frac{\partial Z_7}{\partial t}-2\frac{\partial F}{\partial t}.$$

Because readings at 24-hour intervals eliminate diurnal considerations, we can neglect the last term $\partial F/\partial t$ and write the above equation in the finite form $\Delta Z_5=2\Delta Z_7$ for use over intervals of a day or two. This equation plus the previous temperature relationship is applicable if thickness changes are noted in the same airmass on consecutive days. This equation applied to the changes shown in figure 9 permits one to expect warning of 10° F. or more over Kansas and Oklahoma. A check of the surface temperature changes between the times of figures 3A and 5A shows rises of 10° to 15° F. in that area. Factors contributing to the differences between temperature changes expected over California from the above equation and those which actually occurred include: (a) a change of F over that region due to a change of airmass, (b) variations in cloudiness, (c) precipitation, and (d) wind. The actual changes over California during the period discussed above were limited to slight cooling at the higher elevations.

ACKNOWLEDGMENTS

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